The relationship between tropospheric wave forcing and tropical lower stratospheric water vapor

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The relationship between tropospheric wave forcing and tropical lower stratospheric water vapor

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Abstract

The compact relationship between stratospheric temperatures (as well as ozone) and tropospheric generated planetary wave activity have been widely discussed. Higher wave activity leads to a strengthening of the Brewer-Dobson (BD) circulation, which results in warmer/colder temperatures in the polar/tropical stratosphere. The influence of this wave activity on stratospheric water vapor (WV) is not yet well explored primarily due to lack of high quality long term data sets. Using WV data from HALOE and SAGE II, an anti-correlation between planetary wave driving (here expressed by the mid-latitude eddy heat flux at 50 hPa added from both hemispheres) and tropical lower stratospheric (TLS) WV has been found. This appears to be the most direct manifestation of the inter-annual variability of the known relationship between ascending motion in the tropical stratosphere (due to rising branch of the BD circulation) and the amount of the WV entering into the stratosphere from the tropical tropopause layer. A decrease in planetary wave activity in the mid-nineties is probably responsible for the increasing trends in stratospheric WV until late 1990s. After 2000 a sudden decrease in lower stratospheric WV has been reported and was observed by different satellite instruments such as HALOE, SAGE II and POAM III indicating that the lower stratosphere has become drier since then. This is consistent with a sudden rise in the combined mid-latitude eddy heat flux with nearly equal contribution from both hemispheres. The low water vapor and enhanced strength of the Brewer-Dobson circulation has persisted until now. It is estimated that the strengthening of the BD circulation after 2000 contributed to a 0.7 K cooling in the TLS.

1 Introduction

Stratospheric water vapor (WV) plays an important role in determining radiative and chemical properties of the middle atmosphere. As a primary source of odd hydrogen in the stratosphere, it controls ozone loss through gas phase chemistry. In addition, the

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coupling processes between HO_x and NO_x/ClO_x also affect ozone destruction by other catalytic reaction cycles. It is also an integral part of polar stratospheric clouds which are responsible for enhancing catalytic heterogenous chemical ozone loss reactions. Some observations indicate that the WV volume mixing ratios (VMRs) in the stratosphere have been increasing by about 1% per year since 1981 (Oltmans et al., 2000; SPARC, 2000; IPCC, 2001). Estimated trends in stratospheric WV above 18 km are lower in the satellite data records and are only significant from 1996 until 2000 (IPCC, 2001). Such an increase in stratospheric WV can have serious implication on the future evolution of the ozone layer as it can enhance homogenous as well as heterogenous chemical ozone loss (Shindell et al., 1999; Tabzadeh et al., 2000; Stenke and Grewe, 2005), and can delay ozone recovery (Shindell, 2001).

Initially it was believed that methane oxidation as well as WV directly emitted by aircrafts might have contributed to rising levels of WV (or OH concentration) in the stratosphere. But observed trends are substantially larger than can be attributed to the observed changes in stratospheric methane. Another important source of stratospheric WV is the direct transport through the tropical tropopause that is regulated by changes in tropical tropopause temperatures (Brewer, 1949; Rosenlof, 2003). The tropical tropopause temperatures have actually decreased, and not increased as would be needed for the stratospheric H_2O increase (SPARC, 2000).

Here, we look at another important mechanism that influences stratospheric WV trends by the changes in the Brewer-Dobson (BD) circulation strength. The BD circulation transports most of the chemical species from source region (tropics) to the mid- to high latitude stratosphere. It has a rising branch in the tropics and a downward branch in mid- to high latitude stratosphere. In addition, it influences stratospheric temperatures through adiabatic compression in the polar region and expansion of air in the tropics (Andrews et al., 1987). The BD circulation is primarily driven by breaking of tropospheric generated planetary waves (e.g. Rossby waves) in the mid-latitude stratosphere. The amount of momentum deposited by these waves is generally measured in terms of the Eliassen-Palm (EP) flux. The eddy heat flux which is directly proportional

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to the vertical component of EP flux, is the measure of magnitude of planetary wave driving and the strength of the BD circulation, which has been linked to changes in stratospheric ozone and temperatures at high latitudes (Fusco and Salby, 1999; Newman et al., 2001; Weber et al., 2003).

In this study we show that the inter-annual variability of zonal mean WV VMRs in the tropical lower stratosphere (TLS) are tightly controlled by the inter-annual variability of the strength of the BD circulation that is proportional to the magnitude of wave activity added from both the hemispheres (Sect. 2). The decrease in stratospheric water vapor observed after 2000 (Randel et al., 2004, 2006) in connection with BD circulation changes is discussed in Sect. 3.

2 Planetary wave driving and TLS WV

The seasonal cycle of HALOE TLS WV VMR averaged between 16–20 km altitude and 20° S–20° N together with the global 50 hPa eddy heat flux is shown in Fig. 1. Monthly mean WV VMRs shown are derived from HALOE V19 data (Harries et al., 1996). The water vapor profiles were weighted with the inverse of the squared measurement errors before taking the monthly mean. As shown in earlier studies (SPARC, 2000), WV in the TLS starts decreasing in November and reaches minimum in January-February and starts increasing afterwards with a seasonal cycle of about ±0.5 ppmv. This annual minimum in WV reaches 22 km altitude in May–June depending on the speed of ascending motion (Niwano et al., 2003).

The seasonal variation in WV entry WVMRs in the tropics is primarily due to the upwelling from the tropical tropopause layer (TTL) which influences both temperatures and ascending motion in the TLS (Niwano et al., 2003). TTL temperatures controls the dehydration mechanism while ascending motion controls the amount of WV entering in to the stratosphere. This upwelling is maximum during northern hemispheric winter. This can be explained by a superposition of wave activity (or eddy heat flux) in both hemispheres (Salby et al., 2003). The maximum of the eddy heat flux is observed from November to March during NH winter and in the southern hemisphere (SH) from September to November, so that the total flux from both hemispheres shows large values from September to March (Randel et al., 2002a). This produces a seasonal variation in the TLS temperatures (Yulaeva et al., 1994; Randel et al., 2002b) and it explains the so-called tape recorder effect in tropical stratospheric WV with low (high) WV when tropical tropopause temperatures are minimum (maximum) during the seasonal cycle (Mote et al., 1996). Using trajectory calculations and cold point based dehydration assumption, Füglisthaler et al. (2005) showed very good agreement between modelled and observed stratospheric WV VMRs near TTL. They also showed that the cold point temperatures in tropical tropopause region play an important role in controlling WV VMRs in the TLS. TLS temperatures are also influenced by the Quasi-Biennial Oscillation (QBO) and aerosol loading due to major volcanic eruptions (Baldwin et al., 2001; Randel et al., 2001).

Figure 1 also shows the strong inter-annual variability in TLS H_2O vapor and eddy heat flux. The color coding indicates that during years of high/low planetary wave activity (yellow-red lines/violet-blue lines) TLS WV VMRs are generally lower/higher. This is more emphasised in the scatter plot of tropical JFM WV VMRs and integrated 50 hPa eddy heat flux (averaged over 45°–75° latitudes and integrated from September to February and absolute values added from both hemispheres) that shows a distinct anti-correlation between both quantities. Figure 2 shows both SAGE (black solid) and HALOE (orange light symbols) data.

All SAGE V6.2 water vapor profiles (Thomason et al., 2004) with aerosol extinction coefficients at 1020 nm greater than $2\times10^{-4}\,km^{-1}$ (Taha et al., 2004) have been filtered out, which means that data from 1984–1985 and 1992–1994 are excluded here. HALOE data are shown for all the available years (1992–2005). Lower stratospheric WV VMRs from SAGE are systematically lower than HALOE as also shown by Taha et al. (2004). The JFM WV VMR means were calculated from monthly mean values requiring at least five observed profiles per month. Data gaps for months with less than five profiles were filled by values using harmonic analyses containing annual and semiannual terms. Excluding data from 1997, the correlation between TLS water vapor and eddy heat flux is −0.85 for HALOE (1992–2005) and −0.68 for SAGE II. Without the year 1991, the anti-correlation for SAGE improves to −0.77. As expected higher global wave activity in a given year leads to lower WV VMRs in the tropical stratosphere at the end of the NH winter. The period September-February for eddy heat flux and January-March for WV were selected for the highest anti-correlation between both quantities. It should be kept in mind that the near global sampling of the solar occultation instruments like SAGE II and HALOE requires almost one and a half month. This leads to rather low sampling of the tropical stratosphere for calculating monthly means and may (in parts) influence the correlation between both data sets.

The year 1997 shows very low TLS WV VMR, despite the fact that the winter eddy heat flux is quite low. An explanation for the extreme departure from the linear relationship for both satellite data is not known. In early 1997 the Southern Oscillation Index shows very low values indicating the beginning of a strong El Niño event, however, in JFM of 1998 the TLS WV VMRs appears to be normal, although SOI remained very low.

3 Decrease in TLS H_20 vapor after 2000

As already noticeable in Figs. 1 and 2 and reported in other studies, lower WV VMRs have been observed in the TLS since 2001 (Randel et al., 2004, 2006) indicating that the stratosphere has become drier in recent years. Lower WV VMRs are also found in NH mid- to high latitudes with a time lag of a few months due to isentropic transport in the lowermost stratosphere as confirmed by POAM III data (Randel et al., 2006). Figure 3 shows the WV VMR anomaly time series from HALOE in the tropics and POAM III at NH mid- to high latitudes (Nedoluha et al., 2002) in the lowermost stratosphere. Both data sets show a clear drop in the anomalies after 2000. At the same time a sudden increase in the monthly mean global eddy heat flux is evident as shown in the same figure. The enhancement of the BD circulation as indicated by the increased eddy heat

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flux is consistent with increased tropical upwelling velocities during the same time period and a weak average enhancement in the subtropical convergence of the EP fluxes during the period 2001–2004 in comparison with the late 1990s (Randel et al., 2006). From the global eddy heat flux time series separated by hemispheres (Fig. 3) it can be concluded that each hemisphere contributed roughly equally (NH being slightly larger) to the global anomalies.

The strengthening of the BD circulation in the NH in recent years has been associated with increased Arctic stratospheric winter temperatures and rapid increases in NH total ozone since the middle nineties (Dhomse et al., 2006). An anti-correlation between tropical and mid- to high latitude lower stratospheric temperatures exists on seasonal and inter-annual time scales (Yulaeva et al., 1994; Salby and Callaghan, 2002) so that a corresponding recent cooling of the TLS and/or TTL could be expected. Figure 4 shows a monthly tropical temperature anomaly time series at 70 hPa from NCEP/NCAR reanalysis (hereafter NCEP) and a regression analysis that quantifies the contributions from QBO, solar cycle variability, stratospheric aerosols, global eddy heat flux, and linear trend terms (Dhomse et al., 2006). For each month of the year two QBO terms (50 and 30 hPa, 24 fitting constants), one eddy heat flux term (12 fitting constants, no time lag), and a linear trend term (12 fitting constants) are included. One fitting constant for each major volcanic eruption and solar term are also included (for details on regression see Dhomse et al., 2006). Our regression analysis indicates that the strengthening of the BD circulation contributed to a 0.7 K cooling at 70 hPa (approximately just above the TTL) when comparing a five year period before and after the drop in water vapor anomalies in 2000 (see the horizontal bars in the heat flux contribution (HTF) to the temperature trend in Fig. 4).

The change in the vertical temperature profile from NCEP and the regression analysis between both time periods is shown in Fig. 5. From the NCEP data an apparent temperature change after 2000 is only evident near the TTL (100–70 hPa), but the regression analysis at all pressure levels indicate a nearly constant cooling of about 0.7 K throughout the TLS from the increase in the strength of BD circulation. The cooling trend in the TLS above 70 hPa is partially cancelled by the warming effect from higher solar activity in the later period, but the regression model overestimates the cooling at 50 and 30 hPa. The main reason for this discrepancy is the strong negative temperature anomaly in 1997 with very low global eddy heat flux (normally associated with warm temperature anomalies) which makes the late 1990 period appear warmer in the regression. The linear trend contribution shown here is the average influence from the monthly long-term downward trends for more than two decades. There is some considerable debate on the significance of such trends since changes in radiosonde operations (cooling bias) and changes in the assimilation scheme may strongly impact such long-term trends particularly near the TTL (see discussion in Randel et al. (2004, 2006) and references therein).

As this study focused more on the sudden change after 2000, a more important issue here is the switch from TOVS to ATOVS data in the NCEP reanalysis in July 2001 (Randel et al., 2004). This is illustrated in Fig. 6, which shows temperature anomalies in TLS from various analyses, ECMWF 40 Year Re-analysis or ERA40 (1958–2002), ECMWF operational analysis data set (2001–present), NCEP (1948–present), and Hadley Centre's radiosonde temperature data (HadAT data set). The HadAT data set (http://www.hadobs.org) is an analysis of the global upper air temperature record from 1958 until present based on radiosonde data (Thorne et al., 2005). Figure 6 shows monthly mean temperature anomalies over the tropics (20° S–20° N) at 100 hPa (bottom), 50 hPa (middle) and 30 hPa (top). Anomalies are calculated by subtracting the climatological monthly means from the 1965–1995 period. There are significant differences among these analysis at 100 hPa, that corresponds roughly to the TTL altitude. Large positive temperature anomalies at all three pressure levels are associated with the major volcanic eruptions in 1983 (El Chichon) and 1991 (Mt. Pinatubo). More important here is that after 2000 no significant bias in the NCEP data with respect to the homogenized radiosonde data is seen. This means that the estimated cooling contribution from changes in planetary wave driving appears not to be biased. One should also note that a reduction in the observed ozone near the TTL by about 10%

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also reduces temperatures via radiative feedbacks as shown by Randel et al. (2006). Temperature data combined from ERA40 and ECMWF operational analysis have not been used in the regression analysis primarily due to significant differences between these two data sets. The cold bias in ECMWF operational data ranges between 0.7 K to 2 K in the tropical stratosphere during the overlapping period (January 2001–August 2002).

As model studies have shown that the variability in TLS water vapor are well explained by cold point temperature variability (Füglisthaler et al., 2005) and given the strong link between stratospheric circulation and water vapor as shown by Randel et al. (2006) and in this study, it is somewhat surprising that the 100 hPa tropical temperature anomalies (probably closest to the thermal tropopause) appears to have a more minor contribution from BD circulation strength changes as suggested by Fig. 5. A larger cooling effect comes from the linear long-term downward trend at this level as derived from the regression over 27 year of data. This may however be an indication for the larger uncertainty in the temperature data as mentioned earlier, which may somehow mask the more step like contribution from the BD circulation changes after 2000 at the TTL (~100 hPa).

4 Conclusions

We have shown that satellite data from HALOE and SAGE show interannual variability in TLS WV anomalies that are anti-correlated with the strength of the BD circulation with an exception of few years (1991, 1997) with extreme departure from that relationship which is currently not understood. In addition, this relationship also explains that the increasing trends in stratospheric WV until 2000 are most probably due to decrease in planetary wave driving during that period. There is now clear evidence that the drop in the lower stratospheric WV anomalies after 2000 that has persisted until present can be directly related to raising planetary wave forcing with nearly equal contribution to the increase from both hemispheres. From a regression analysis of tropical temperature data a cooling of about 0.7 K in TLS was estimated in connection with the recent enhancement in BD circulation strength. A coherent picture of changes in mid- to high latitude ozone (Dhomse et al., 2006) and lower stratospheric WV in connection with BD circulation changes emerges that may have important implications in a future changing climate.

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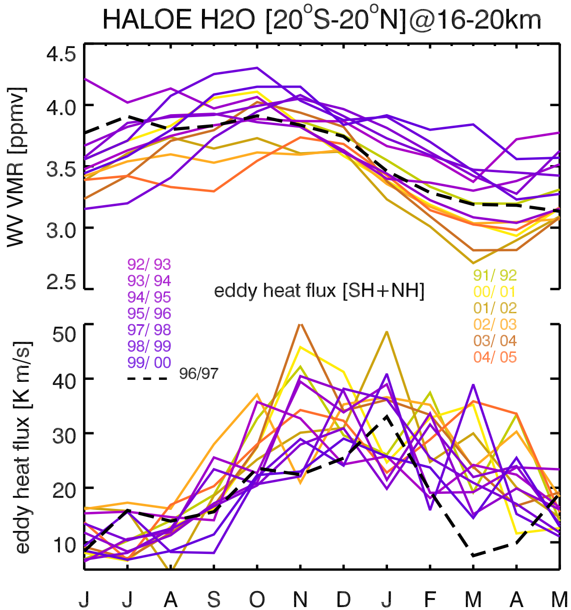

Fig. 1. Annual cycle of monthly mean tropical WV VMRs averaged averaged between 16 and 20 km and 20° S–20° N from HALOE V19 data (top) and monthly mean mid-latitude (45°–75°) eddy heat flux at 50 hPa from added from both hemispheres. HALOE V19 data is obtained from http://haloedata.larc.nasa.gov/, and is avaliable from June 1991 until November 2005 (Kalnay et al., 1996).

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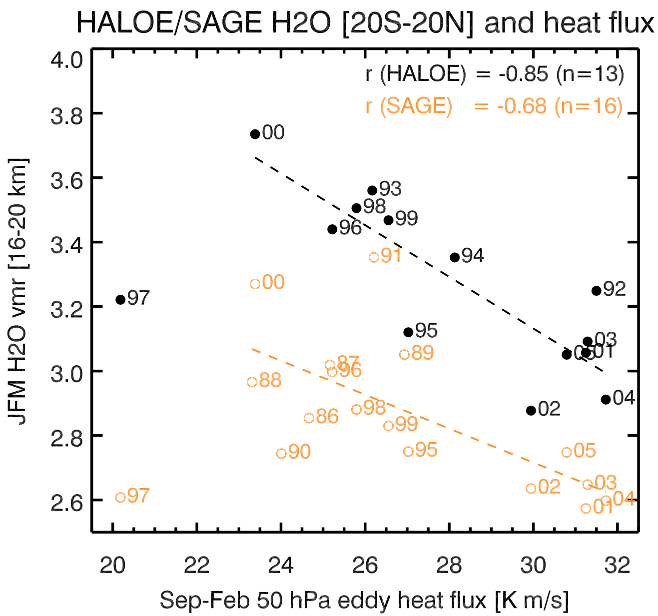

Fig. 2. Anti-correlation between JFM TLS WV VMRs (16–20 km, 20° S–20° N) from SAGE (open orange circles) and HALOE (filled black circles) and wintertime 50 hPa eddy heat flux (September–February and added from both hemispheres). The year 1997 shown here has not been included in correlation calculation. SAGE V6.2 data (October 1984–August 2005) is obtained from ftp://ftp-rab.larc.nasa.gov/pub/sage2/v6.20/.

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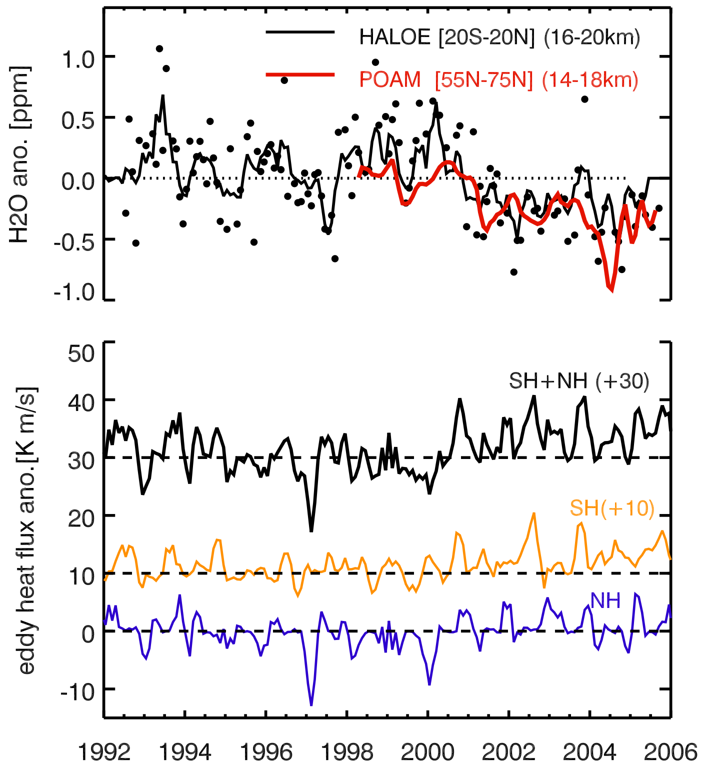

Fig. 3. Top panel: monthly mean H_2O vapor anomalies from HALOE (16–20 km, 20° S–20° N) in the tropics (solid) and POAM III (14–18 km, dashed) at middle to high NH latitudes. Both lines are three month mean WV VMRs, while circles are monthly mean HALOE values (Update from Randel et al., 2006). Bottom panel: Time series of monthly mean eddy heat flux (50 hPa) from each hemisphere and globally. POAM III data were obtained from http://www.cpi.com/.

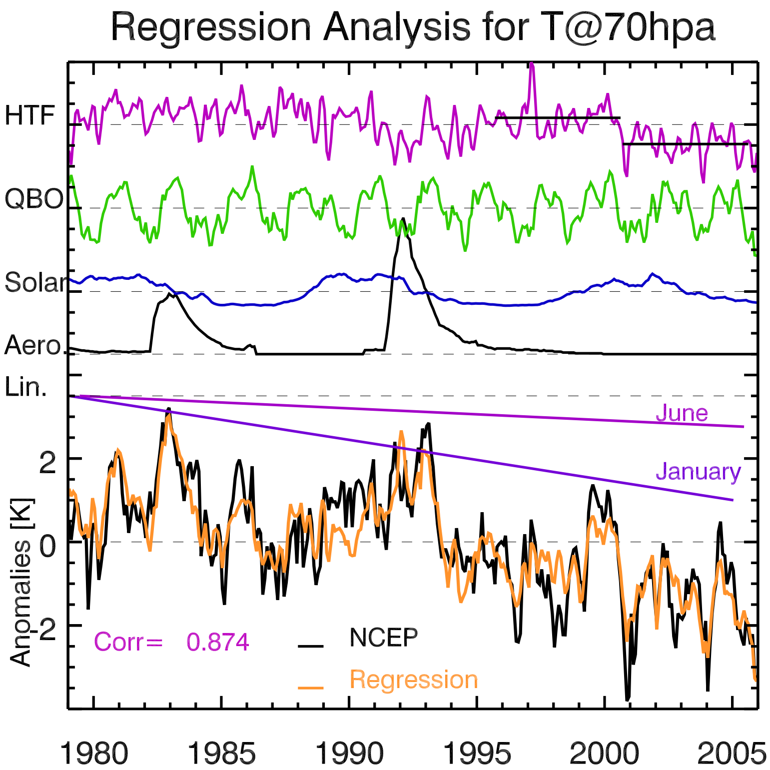

Fig. 4. 70 hPa temperature anomalies from NCEP and regression model (15° S–15° N) with contribution from QBO, linear trends, solar variability, stratospheric aerosols, and global eddy heat flux (HTF). Linear trends are only shown for minimum (June) and maximum (January) downward trends. The strengthening of the BD circulation after 2000 resulted in a cooling of about 0.7 K at this pressure level compared to the late 1990s (see horizontal bars in the HTF panel and Fig. 5). For details on regression analysis see Dhomse et al. (2006).

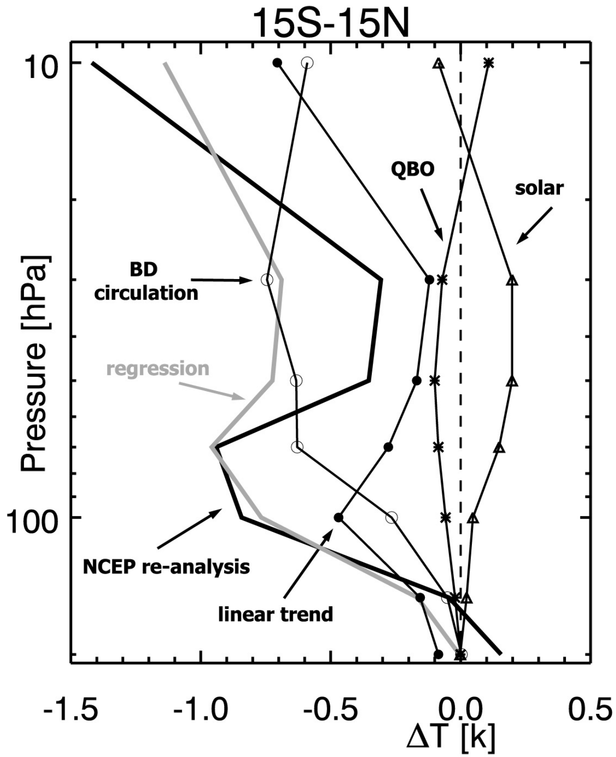

Fig. 5. Average temperature profile change between the periods July 2000–June 2005 and July 1996–June 2000. Thick solid lines: NCEP re-analysis (black) and regression analysis (grey). Other lines: individual contributions from the regression as indicated. Note that the regression coefficients were derived from 27 years of data (1979–2005) as shown in Fig. 4.

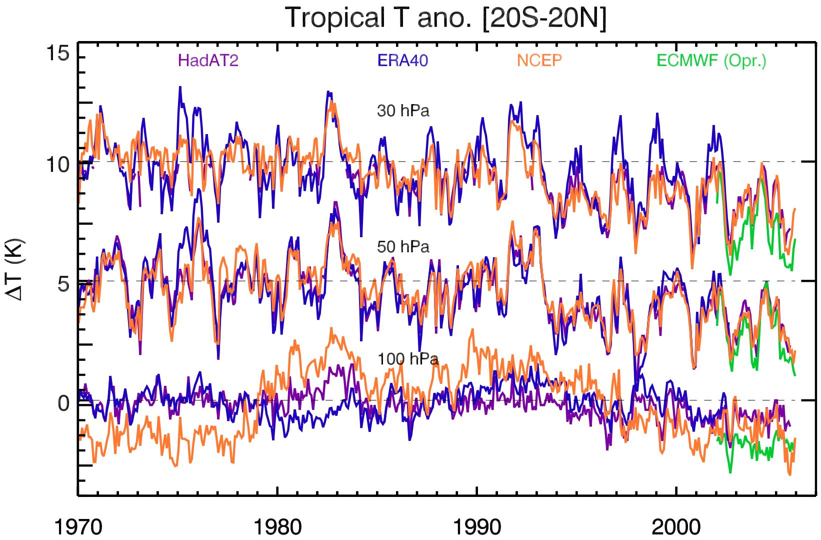

Fig. 6. Temperature anomalies in the TLS (20° S–20° N) from different meteorological analyses. Temperature anomalies from ERA40 (Uppala et al., 2005) are shown in blue and from ECMWF operational data (http://www.ecwmf.int) are shown in green. NCEP (Kalnay et al., 1996) and HadAT2 (Thorne et al., 2005) are shown in violet and orange respectively. Monthly mean anomalies are calculated using climatological mean values for 1965–1995. Temperature anomalies at 50 hPa and 30 hPa are shifted by 5 K and 10 K respectively. Also note that, temperature anomalies from ECMWF operational data show a cold bias compared to ERA40. Using twenty months of overlapping period (January 2001 till August 2002), observed cold bias in operational data is of the order of 1.1 K at 100 hPa, 0.8 K at 50 hPa and 1.8 K at 30 hPa.